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REGIONALIZED TEMPERATURE VARIATIONS IN THE UPPER 400" KM OF THE EARTH'S MANTLE

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ABSTRACT

Tectonically regionalized variations in the tempeı ature of the upper 400 km of the Earth's mantle are estimated from analysis of global seismic Lmvcl-time data catalogued by the International Seismological Centre (ISC). Seismic parameter profiles are determined from estimates of P and S velocities obtained by tau inversion. Summary phase diagrams for the olivine and pyroxene-garnet subsystems are constructed in conjunction with a thermodynamic potential formulation that allows self-consistent determination of density, bulk modulus and adiabats throughout the pressure and temperature regimes of the mantle. Perturbations in estimated seismic parameters are expressed in terms of variations in temperature using the model temperature derivatives of the bulk modul us and density at a given temperature and pressure. Confidence bounds on the velocity estimates are used to place corresponding bounds on the constructed seismic parameters. A simple differential relationship is solved iteratively to obtain a temperature variation for a given variation in seismic parameter. This approach allows the cstinlation of a range of seismicall y/determined temperature variations by employing a given compositional model. Results indicate that while the P and S velocity variations in the upper mantle are consistent with the tectonic regionalization, variations in V_p/V_s ratios are irregular. This leads to unstable estimates of the seismic paraı neters and thus estimates of mean temperature anomalies, typically within 600°C of the weighted mean, that are inconsistent with the region alized seismic data. A comparison of two compositional models is used to show the trade-off with estimated temperature variations. A refined regionalization and analysis of a larger ISC data set are suggested in order to stabilize the S velocity inversion, reduce statistical uncertainties on the seismic parameters, and thus improve constraints on estimated temperature variations.

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1. INTRODUCTION

Seismic velocities are governed by the temperature and pressure dependence of the composition, chemistry and mineralogy of the Earth's interior. Inversion of global seismic travel-time dots thereby is a means of constraining mantle composition, inferring differences in mantle state, anti comparing calorimetric and thermoelastic models [e.g. Bina and Wood, 1987; Ita and Stixrude, 1992].

In this study, compressional (P) and shear (S) velocities arc estimated by tau inversion of 1 sec body wave travel-time data cataloged in the Bulletin of the International Seismological Centre (1 SC). Lateral heterogeneity is parameterized by P and S tau-slowness estimates appropriate for oceanic and continental regions following the tectonic regionalization model of Jordan [1981]. Regionalized tau inversion lacks the three-dimensional appeal of tomographic imaging. However, the tau method is quantitative and statistically sound. The analysis dots not rely on merging compressional and shear velocities obtained from different data sets, frequency responses and geographic regions, but instead uses travd-times in a manner that is parametrically consistent for global P and S data. Joint analysis of P and S velocities not only helps delineate the depth extent of differentiation between oceanic and continental regions, but helps infer the compositional, chemical and thermal nature of the mantle given the particular response of each velocity mode to the lateral heterogeneity. This is especially the case in the uppermost mantle, where the observed lateral heterogeneity is most consistent with the tectonic regionalization.

From the P anti S velocity estimates, seismic parameter profiles arc constructed, defined as the ratio of bulk modulus to density [see Birch, 1952]. Confidence bounds on the velocities obtained by tau inversion arc used to place bounds on the seismic parameters. Piclogitic and pyrolitic models [Ita and Stixrude, 1992] arc used to yield temperature derivatives of the bulk modulus anti density at a given temperature and pressure in order to estimate a range of seismically-determined temperature anomalies from the variations in seismic parameter. While the shear velocity structure is particularly sensitive to mantle temperature and solidus, the model temperature derivatives of the shear modulus are poorly known. '1'bus, using the seismic parameter is a convenient means of estimating thermal anomalies without sensitivity to uncertainties in modeling the shear modulus.

2. SEISMIC DATA ANALYSIS

Determination of tectonically regionalized velocity-(icplh functions from estimates of tau-slowness functions is described by Tralli and Johnson [1986 a]. Over 1.25 million P ray paths with sour-cc depths less than 70 km were obtained from the 1 SC Bulletin over a period spanning 7.5 yrs. However, this data set yielded only about 150,600 corresponding S ray paths, but with a fractional distribution per region comparable to that of the P data. The small size and the larger scatter in the S travd-lime data lead to greater numerical sensitivity and uncertainties in the S velocity inversion, and thus increased uncertainties in the constructed seismic parameters. This, in turn, decreases the ability to constrain estimates of regionalized temperature variations.

The global tectonic regionalization of Jordan [1981] is adopted, which consists of a 5 by 5 degree surface gridding, with the inclusion of oceanic trenches based on anomalous reduced travel-tirnc distributions observed in the more robust P data set [see Tralli and Johnson, 1986a]. Briefly, for regionalized tau inversion, seismic ray paths tire grouped according to the tectonic regions associated with the source, receiver station and surface projection of the ray turning point. A simple statistical averaging then allows construction of tau estimates appropriate for each region type, with slowness-dependent source and receiver perturbations explicitly formulated [Tralli and Johnson, 1986 b]. These "single region" equivalent tau estimates are inverted independently to estimate P [Tralli and Johnson, 1986a] and S (this study) velocities throughout the entire mantle.

The tau analysis for S travel-time data follows that developed for P data, except that selection criteria for tau estimates (see Tralli and Johnson, 1986a) arc less stringent given the larger variances in the S data. Furthermore, the slowness (p) and epicentral distance (Δ) intervals used to reduce the raw S travel-times are obtained by linearly fitting data in 1 to 10 deg sliding windows over 0.10 deg increments from 10 to 85 deg in epicentral distance. The slopes of the fits then are rounded off to the nearest of 35 slowness values selected *a priori* for the entire mantle, while selecting a respective set of consecutive A intervals for each of the seven tectonic region types. Thus, data from each region arc reduced with common slowness values, but each slowness may correspond to a different epicentral distance interval, thereby allowing the manifestation in the data reduction of velocity differences beneath the various tectonic regions. This method of choosing p-A intervals for the entire epicentral range is more efficient than the algorithm used to empirically segregate the upper mantle. travd-time data in the P study, and more general than using common p-A intervals, as for the lower mantle in the P study.

2.1 Regionalized Velocity Inversion

Lateral variations in P and S velocities are expressed as differences about a mean obtained by weighing the various velocity profiles according to the areal surface (Table 1) reprcsented by each tectonic region. This weighting adjusts, to some extent, the over- and under-sampling of certain regions given the geographic distribution of seismicity and seismographic stations [Tralli and Johnson, 1986a].

To ensure that the estimated weighted mean velocity profiles are reasonable and without any potential bias, they are compared to the IASP91 model [Kennett and Engdahl, 1991] (see Fig. 8). Discontinuities, such as at 41 () and 660 km depth, are smoothed out in the velocity inversion since travel-time triplications in the ISC data tire not sufficiently resolvable for tau inversion. This smoothing out of the discontinuities unfortunately limits the ability of this study to infer temperature anomalies to only the upper 400 km of the mantle.

The P velocity anal ysis of Tralli and Johnson [1986a] points out significant differences between oceanic and continental regions to a depth of 700 km. Predominant features are the gradient in the velocity residual from the regionally weighted mean below oceanic regions, whereas continental platforms and shields show compensation in the sign of the velocity residual between 350 and 700 km. Figs. 1-7b show differences in each S velocity profile from the weighted mean and can be compared with the corresponding P velocity variations in Figs. 1 -7a [from Figure 4 of Tralli and Johnson, 1986a]. Such a comparison indicates a factor of two between the magnitude of P and S velocity variations from their respective regionally weighted means.

Table 1

Tectonic Regions and Fractional Surface Area

1	Young oceans (< 25my)	().13
2	Intermediate-age oceans (2,5-1()() my)	().34
3	Old oceans (> 100rny)	().13
4	Active continents (orogenic zones, magmatic bolts)	().19
5	Continental platforms	().1()
6	Continental shields	().07
7	Oceanic trenches	().04
		1.0()

There is a trend of increasing surface S velocities (at 33 km depth after a crusts] correction is applied [see Tralli and Johnson, 1986a]) from 4.32 km/s in region 1, and 4.47 km/s in region 2, to 4.53 km/s in region 3. (Tectonic region indices are listed in 'J'able 1, and correspond to figure numbers). This trend agrees with the I' velocity results, and is consistent with the square root of crustal age dependence of the oceanic regionalization [Jordan, 1981]. A similar increasing trend is noted in the continental regions, with surface velocities progressing from 4.23 km/s (region 4) to 4.45 km/s (region 5) and 4.62 km/s (region 6).

Region 4 suggests the greater sensitivity of shear waves to zones of higher temperature, particularly in the upper '250 km (Fig. 4b). Residual velocities of regions 5 and 6 are entirely positive in the upper 400 km (mor eso and significantly at the 99.9% confidence level for region 6). Platforms indicate consistently positive residual velocities, with variations from the weighted mean of 1.6 to 2.2% extending from the surface to about 280 km in depth. The compensation seen in the P velocity variations beneath platforms is only hinted at in the S results. The residual S velocity beneath shields is significant at the 99.9% confidence level to a depth of about 360 km, with a 3.5% residual velocity variation from the mean.

Oceanic trenches (region 7) are significantly slower in S velocity to about 280 km in depth, below which there is a tendency towards compensation in the sign of the velocity variation that extends below 400 km. The P velocity in the shallow mantle beneath trenches dots not indicate this. As in active continental regions, the S velocities beneath trenches may be suggestive of shear softening and partial melting of subducting slab material. However, there is insufficient resolution in the inversion and particular] y in the regionalization of trenches at depth [see Tralli and Johnson, 1986a].

The estimated P and S velocity variations are most compatible with the tectonic regionalization in the uppermost mantle. This corroborates the results of other studies [e.g. Hara and Geller, 1994; Su *et al.*, 1994; Nolet *et* al., 1994]. For the purpose of inferring temperature variations, only results in the upper 400" km are considered, for the reason cited earlier regarding the lack of resolution of the tau method as used through mantle transition zone discontinuities. It is evident at this point, however, that P and S velocity variations are not necessaril y similar in form throughout the mantle. This has implications for estimating temperature variations from a single compositional model for all regions, as will bc discussed later given the observed nature of the V_p/V_s ratios and the constructed seismic parameters.

2.2 REGIONALIZED SEISMIC PARAMETER

Regionalized profiles of the seismic parameter, ϕ, are determined from the 1' and S velocities in a straightforward manner from the relation

$$\phi = V_p^2 - 4/3\ V_s^2 \tag{1a}$$

Each velocity profile first is interpolated to common depths at 20 km intervals, since the depth corresponding to a given velocity estimate is determined by the tau inversion and cannot be set *a priori*. There is a slight bias therefore introduced in the interpolation of velocity estimates and their confidence hounds. The seismic parameter is simply the ratio of the bulk compressibility to density, or the squared bulk sound velocity,

$$\phi = K/p \ . \tag{1b}$$

Fig. 8 compares the bulk sound velocity in the upper mantle with IASP91[Kennett and Engdahl, 1991]. Again, there is agreement with the model except at discontinuities.

Variations in regionalized seismic parameters from their weighted mean are shown in Figs. 1 -7c (figure numbers correspond to region indices). The only significant variations in oceanic regions are (positive) between 260 and 400 km beneath region 2 and (negative) from about 150 to 300" km beneath region 3. Continental and oceanic trench regions indicate a trend of decreasing values proceeding from region 4 to region 6. A reversal in the residual seismic parameter anomaly from 350-375 km to 4(H) km in depth is evident beneath all regions, including oceanic trenches.

The discrepancy in the behavior of seismic parameter variations within oceanic regions (namely regions 1 and 2) compared to that within continental regions is suggestive of an inconsistency between the 1'and S velocity estimates. For example', S velocities may be too low beneath active continental regions and too high beneath platforms. However, both of these regions indicate that the S velocity variations are significant at the 99.9% confidence level (see Figs. 4b and 5b). To check the in(cr-consistency of the P and S velocity estimates, profiles of V_p/V_s were determined. Lateral variations in V_p/V_s about their regionally weighted mean are shown (without confidence bounds) in Figs. 1 -7d. Further discussion is deferred to Section 4.

3. ESTIMATION OF TEMPERATURE VARIATIONS

once. variations in regionalized seismic parameters are determined, attention is turned to estimating temperature variations for a given compositional model. Summary phase diagrams for the olivine and pyroxene-garnet subsystems are constructed. A sell-c~)nsiskm[thermodynamic model then is used to determine the density, bulk modulus, and adiabats throughout the pressure and temperature regime of the earth's mantle assuming piclogitic and pyrolitic compositions [see Ita and Stixrude, 1992; 1993 for details].

Perturbations in estimated seismic parameter (ϕ) due to variations in temperature (T) are expressed using model temperature derivatives of bulk modulus (K) and density (p) at a given temperature and pressure according to

$$\phi - \phi_0 = \frac{1}{\rho}\left[\frac{dK}{dT} - \phi_0 \frac{d\rho}{dT}\right](T - T_0) \qquad (2)$$

The subscript O refers to the quantities derived from the thermodynamic potential. This differential relationship is solved iteratively to obtain a temperature anomaly for a given anomaly in seismic parameter. Convergence is obtained when $\phi - \phi_0$ is within 0.03, which occurs typically in less than five iterations with a starting surface adiabatic temperature of 1800°K. An iterative method is needed because the relationship between temperature and velocity is non-linear. Also, note that the 1800°K adiabat is only for the starting point, with final temperatures depending solely upon the velocity profiles and mineralogical model chosen. Thus, temperature variations are not references to the 18()( )°K adiabat, but deviations from the weighted mean temperature profile obtained from the iterative procedure.

Rather than adopt the 99.9% confidence bounds on the P and S velocities, only the 1 σ standard deviations are used to determine corresponding bounds on ϕ that yield significant variations from the weighted mean, albeit with reduced confidence. The lower and upper bounds at a given depth are not symmetric about the mean due to the nature of the tau inversion. The error bounds on ϕ then are inverted to yield bounds on the corresponding estimates of (mean) regionalized temperature variations.

4. Discussion of Results

Near-surface source and receiver corrections are determined from the regionalized tau method and presumably remove systematic effects that otherwise would be mapped into lateral variations in upper mantle velocity [see Tralli and Johnson, 19Xotl]. However, it is possible that the seismic ray path turning point sampling per region in the upper 400 km is not representative of the overall distribution of regionalized data throughout the entire mantle. The Weighting scheme adopted thus would not account for sampling biases as intended. The changes introduced hy using the fractional percentage of total ray paths for each region (for example, see '1'able 1 of Tralli and Johnson [1986a]) rather than the surface area values derived from the regionalization model are not significant, and do not account for peculiarities in comparing the P and S behavior and the noted discrepancies with the tectonic regionalization.

Variations in temperatu re about a regionally weighted mean are shown in Figs. 1-7c using a pyrolitic compositional model [Its and Stixrude, 1992]. Young oceanic regions indicate only two depths at which temperature variations are significant at the 1 σ level, namely at the surface and at about 335 km, where estimated temperatures are 400°K higher and 600°K lower than the mean, respectively. For this and all other regions, anomalies at depths approaching 400 km may be due to the lack of resolution in the velocities estimated by the tau inversion as used, and thus may be artifacts of resultant smoothing through transition zones. Intermediate-age oceans are significantly warmer than the mean by about 100°K from the surface to a depth of about 100 km, and show a similar negative variation at a depth of about 335 km. Older oceans indicate a significant positive temperature variation of approximately 200°K at about 130 km to 600°K at about 300 km depth. Any suggestion of warming to cooling temperature progressing from young to old oceanic regions occurs above about 110 km in depth. Older oceans in fact are relatively warmer at depths centeredabout *200"* km.

The temperature variations beneath active continental regions using a pyrolitic model are consistent and significant, about -300°K, extending from the surface to a depth of about 260 km, with a reversal to $+300^{\circ}$K defined at a depth of about 335 km. Platforms show the same overall trend; the temperature variations are slightly greater but insignificant relative to the weighted mean. Continental shields indicate a positive and significant temperature anomaly of about $+200^{\circ}$K from the surface to a depth Of about 185 km. A large anomaly of $+1200^{\circ}$K at about 350 km in depth, defined from 270 to 400 km, is indicated and reminiscent of a similar feature beneath regions 4 and 5. Oceanic trench regions indicate a temperature anomaly of -300°K extending from the surface to a depth of about 260 km. An increase in the estimated temperature variation to approximately $+400^{\circ}$K at 335 km depth is marginally significant at the 1σ level but mimics the results beneath all continental regions. Based on the tectonic regionalization, one would expect a relative cooling trend progressing from active to more stable continental regions. Figs. 4c-6c indicate the opposite trend.

Since dK/dT and $d\rho/dT$ are negative quantities throughout the upper 400 km of the mantle, it follows from Eqn. 2 that positive anomalies in seismic parameter correspond to negative temperature anomalies and vice versa, for a given compositional model. However, as noted in Section 2.2, the behavior of the residual seismic parameter variations from young to older continental regions is predominantly contrary to that indicated within oceanic regions (Figs. 1 -7c). For example beneath region 6, both P and S velocities are greater **then** their weighted means, but the bulk sound velocity is less than its corresponding weighted mean (see Figs. 6a-c). This may be attributed to S velocity estimates that are increasingly too large with continental age, accepting the P results given the robustness of that data set. It is also possible that some of the inconsistency in the estimated seismic parameter and temperature variations may be due to lateral variations in composition or volatile content [Anderson, 1994]. This contention is supported by the variation in the V_p/V_s ratios for region 6. The ratio is anomalously low throughout the depth range considered (Fig. 6d), indicating an inconsistency between estimated P and S velocities.

In order to test the sensitivity of temperature anomalies to the chosen mantle composition, we calculate variations in temperature using a piclogitic composition [Its and Stixrude, 1992] (Figs. 1 -7f). Figs. 4f and 7f indicate a lack of convergence in the uppermost depth node due to large differences between the model and constructed seismic parameters. Predicted velocities using this composition are systematically lower than those generated with the pyrolitic model. This leads to lower temperatures, generally by 250 to 350°C, in every tectonic region. "1'bus, the inverse relationship between the perturbations in the P and S velocities and the seismic parameter noted earlier for region 6, for example, may signal a change in mantle composition without a significant variation in temperature.

The relative temperature differences between the various regions arc maintained with either mantle composition model. These relative differences are contrary to what would be expected from the tectonic regionalization. The estimated velocity variations, on the other hand, are consistent with the regionalization individually for P and S. irregularities arise when the velocities are combined to construct regionalized seismic parameters. As noted, the form (magnitude and sign) of the P velocity variations arc not necessarily mimicked by the S variations estimated in this study. in addition to using a single composition from which to derive model parameters for ail regions, the S results thus may be statistically suspect due to a smaller data set and application of a regionalization model that was refined for tau estimation only in the process of analyzing the more robust P data.

5. CONCLUSION

The use 01 variations in the seismic parameter to estimate corresponding temperature variations is straightforward theoretically and provides statistical rigor through the use of P and S velocity uncertainties yielded by tau inversion. Working with the seismic parameter eliminates the need for model temperature derivatives of shear modulus, which are poorly constrained by laboratory work. However, obtaining estimates of temperature variations from variations in seismic velocities is not an easy task in practice. in addition to assumptions regarding mantle composition, it also suffers from the inability of the seismic data to resolve the finer structure of the mantle through travel-time analysis, and from the limitations of the adopted global tectonic regionalization and its application for parameterizing lateral heterogenei t y at depth.

Nonetheless, this study has gone full circle, presenting a formal and quantitative means of estimating mantle temperature anomalies with regionalized ISC trilvcl-time data. Variations in mean regionalized temperature typically proved within 600° 01 the weighted mean profile. The trade-off between assumed composition and estimated temperature variations was shown using piclogitic and pyrolitic models. This indicated that absolute diffe rences in temperature of up to 350° could be explained by a than.gc in composition. Estimated temperature variations are inconsistent with the tectonic regionalization.

A quadratic programming or linear inequality constraints approach could be undertaken to solve for temperature variations corresponding to seismic parameter profiles that lie between the given uncertainties. However, this would be more involved numerically and is not warranted by the resolution of the seismic parameter profiles constructed from these velocity estimates. A joint inversion of P and S velocities also could be undertaken in a manner that constrains V_p/V_s ratios, for example, and therefore improves the estimation of temperat urc anomal ies from the scism i c parameter.

Finally, a refined tectonic regionalization, with smaller surface grid size to perhaps 1° cells, and a larger travel-time data set, recognizing that over ten years of additional ISC data are now available, are suggested in order to check consistency with the regionalization model *a priori*, stabilize the S velocity inversion and (bus improve constraints on temperature anomalies using the statistical uncertainties on the seismic parameter for a given compositional model.

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FIGURE CAPTIONS

Figs. 1-7: Numbers refer to tectonic region indices according, to Table 1. Outages reflect spurious estimates or estimates that exceed bounds on given figure. Dashed lines represent 1 σ error bounds as discussed in text.

a. Variations in regionalized seismic P velocities from the regionally weighted mean;

b. Variations in regionalized seismic S velocities from the regionally weighted mean;

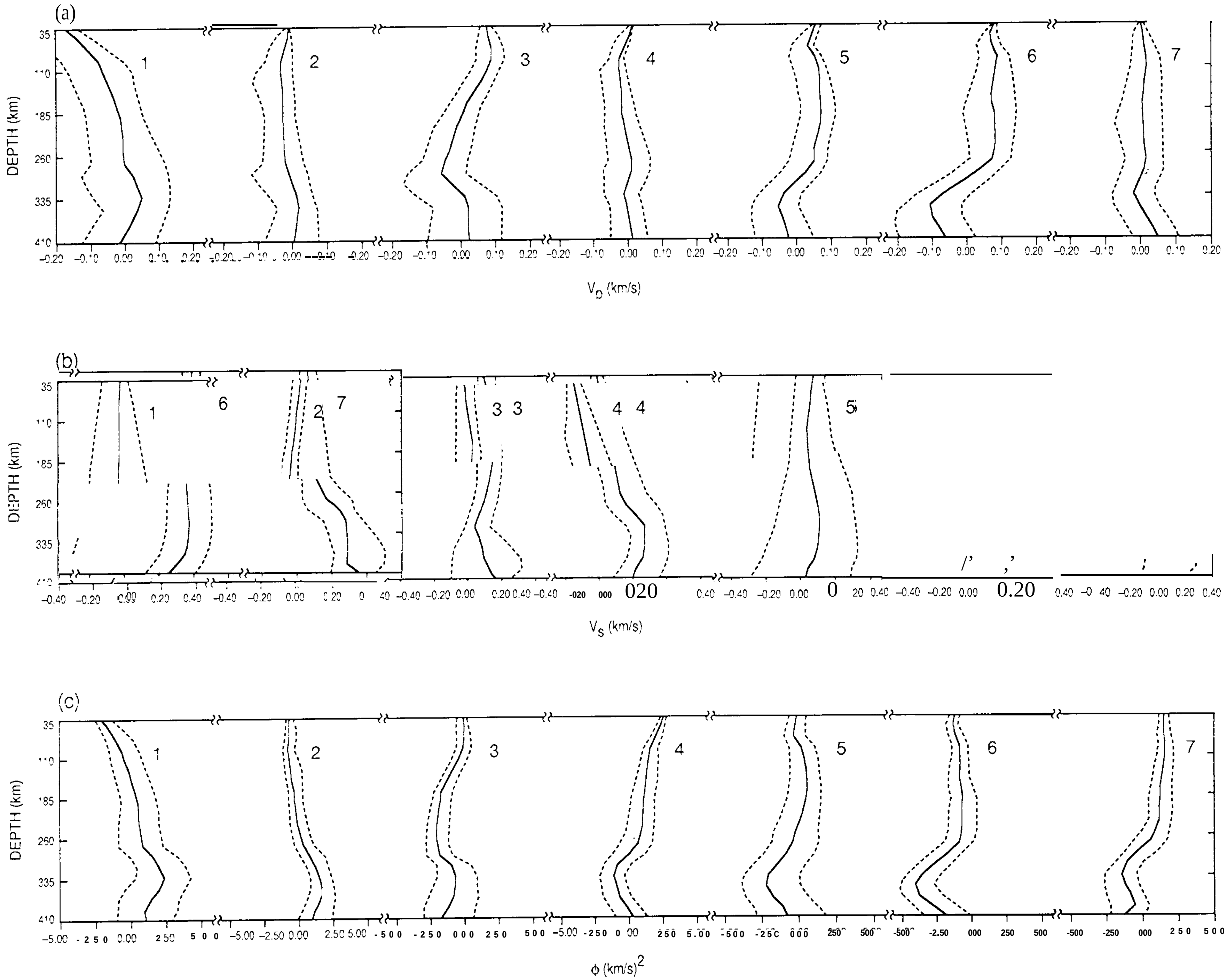
c. Variations in regionalized seismic parameter (ϕ) from the regionally weighted mean;

d. Variations in regionalized V_P/V_S from the regional l y weighted r can;

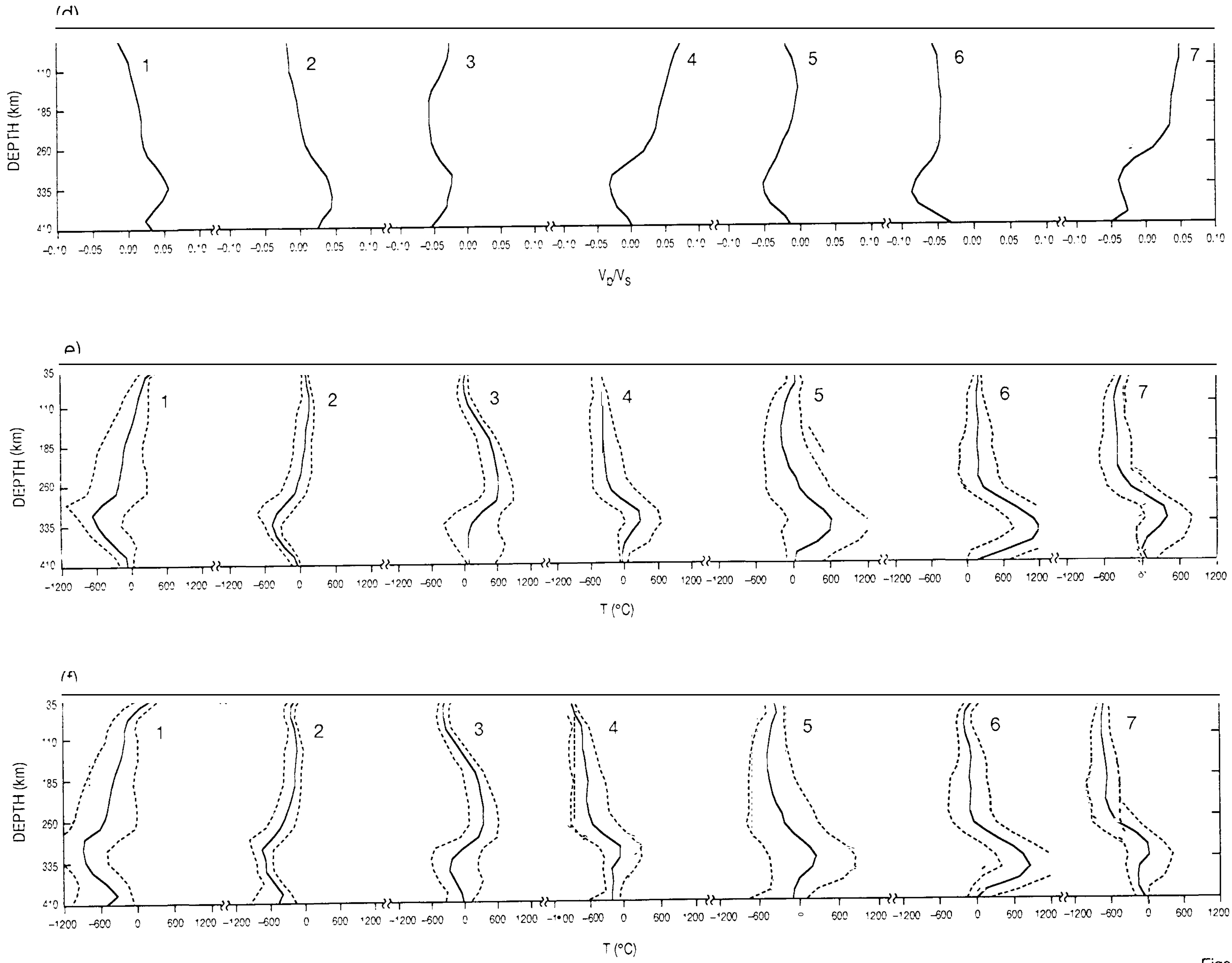
c. Variations in estimates 0! regionalized temperature variations (^{o}C) using a pyrolitic composition [Its and Stixrude, 1992].

f. Variations in estimates of regionalized temperature variations (^{o}C) using a piclogitic composition [Its and Stixrude, 1992].

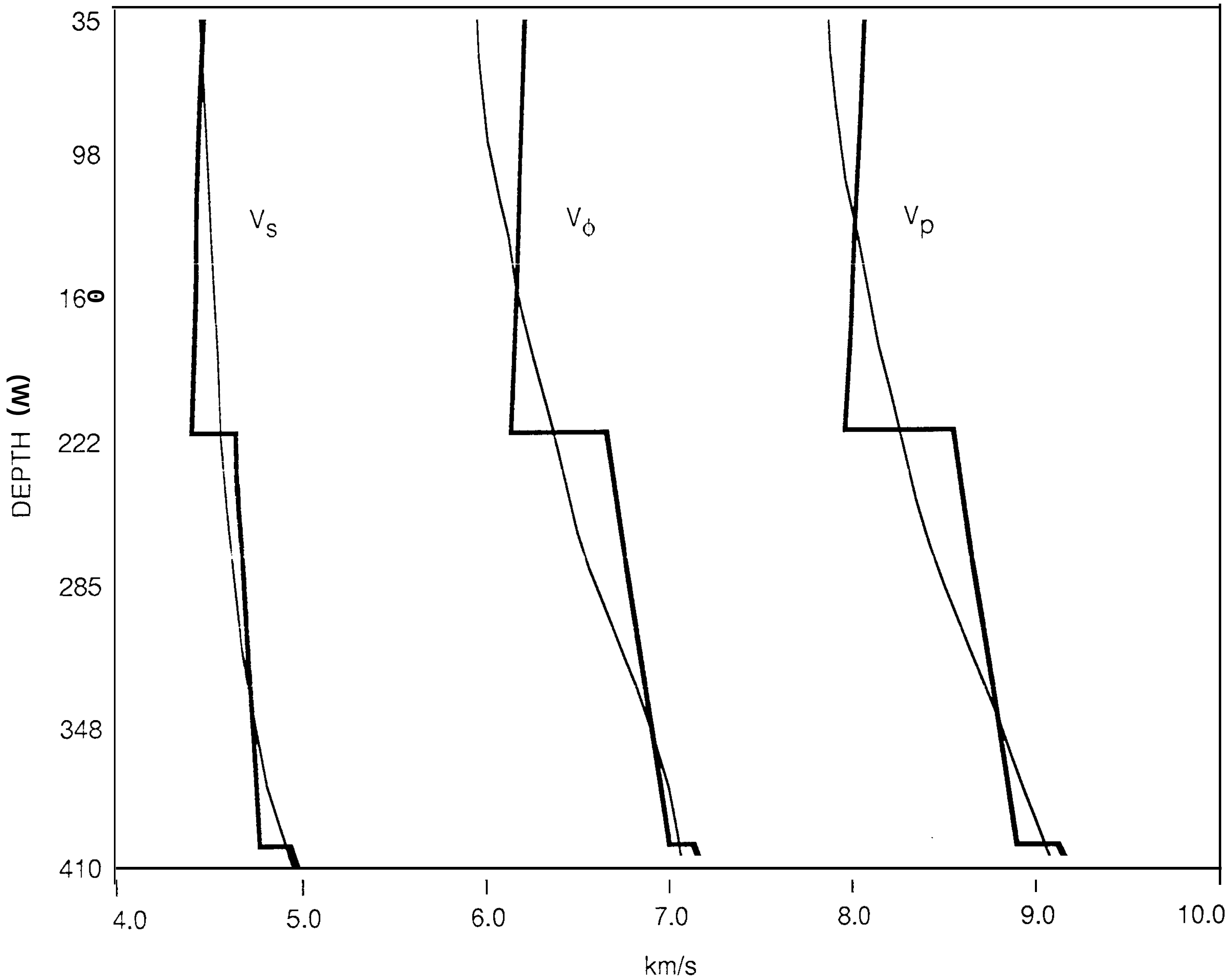
Fig. 8. Comparison of regionally weighted mean V_P, Vs, and bulk sound velocities (V_ϕ) with IASP91[Kennett and Engdahl, 1 991] (in bold).



Figs. 1-7



Figs 7



DEPTH (M)
35
98
160
222
285
348
410
4.0
5.0
6.0
7.0
8.0
9.0
10.0
km/s
V_s
V_ϕ
V_p

Fig. 8